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GEOPHYSICAL MODELLING OF

PLATE TECTONIC PROCESSES

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Equations found in Chapter 1 are taken from Geodynamics. Applications of Continuum Physics To Geological Problems. by Turcotte & Schubert. Chs. 1.2.4.7. Example solutions to equations 1-1 through 1-8 are given in Appendix A.

Chapter I

MANTLE PROCESSES

OVERVIEW

The basic hypothesis of plate tectonics and major earth interactions was given by Jason Morgan in 1968. Stemming from this hypothesis, the most detailed studies done by geophysicists and geologists alike seeking to explain the large-scale plate motions observed in nature have their basis in some form of heat driven thermal mantle convection. The objectives of this paper is not to refute or advocate any specific convective process. Instead, I will attempt to review convection-based phenomena and present a clear and concise analytical summary of its most important geological aspects. The idea of rigid lithospheric plates in relative motion with respect to one another is an important consequence of thermal convection found in the earth's mantle which extends to an approximate depth of 2880km. The actual details of mantle convection is still controversial but the general nature of thermal convection itself is well understood. Thermal convection can occur if a fluid is heated from below, or from within, and is cooled from above in the presence of a gravitational field. The viscous mantle becomes unstable as the hot mantle rocks at depth interact with the colder more dense rocks of the overlying lithosphere. The result is thermal convection in which the colder, heavier rocks descend into the mantle and hotter, lighter rocks ascend toward the surface. The difference in buoyancy between the hot and cold viscous mantle is the driving force of the motion.

A firm command of mantle processes is unrealistic because of the inaccessibility of the mantle region. The primary objective of a geophysicist is to explore the convective engines in the mantle in order to explain how they operate, how they interact with other geologic systems and how they control near-surface processes. It is here where geophysical methods of numerical simulations and modeling can provide new insight into large-scale motions of the mantle and of thermal convection itself.

STATISTICAL ANALYSIS

Density and pressure are key components of the thermal convection process and seem logical as the introductory expressions to be modelled in explaining convective behavior. The pressure (p) of rock in the mantle is given by the hydrostatic equation:

$$(p = \rho\ g\ y) \qquad (1\text{-}1)$$

wherein the following notation is used:

ρ = *density of the mantle rock*
g = *acceleration of gravity*
y = *depth*

Using equation 1-1, a good approximation of the pressure as a function of depth can be calculated for the mantle within the parameters given. Equation 1-1 is a very simplistic description of mantle pressures but for all intense purposes will enhance the basic understanding of thermal convection located in the mantle and in addition, will allow for a more broad analysis of mantle and convective behavior. An exhaustive review of the many elements that may alter the values of the equations given will not be given. I discuss only those factors that will help to produce a general yet thorough modeling of the earth's tectonic processes.

The density of the mantle region for example at 100km can easily be calculated by the rearrangement of equation 1-1. But density variations caused by temperature and pressure changes with depth cannot be overlooked. Material in the mantle and that assimilated from crustal regions is heated from below and from within by the core and from the mantle itself. The heat that is produced originates from two main areas: ancient differentiation during accretion of the earth and the long term radioactive decay of major long-lived isotopes. These two processes generate the majority of the heat found in the interior of the earth. The emphasis on temperature is founded in the fact that heat can be converted to directed motion by the thermal convection process. As mantle material is heated, its density generally decreases because of thermal expansion. With the addition of another parameter, the density can now be calculated through temperature changes using equation 1-2:

$$\rho' = -\rho_o \alpha_v (T_o - T) \qquad (1\text{-}2)$$

wherein the following notation is used:

ρ_o = *reference density (where $\rho' \ll \rho_o$)*
α_v = *volumetric coefficient of thermal expansion (the fractional ϵ volume with temperature at constant pressure)*
T = *temperature of mantle depth of interest*
T_o = *reference temperature (~ surface temperature)*

Equation 1-2 allows the density to be modelled in the mantle for which direct measurement is thus far impossible. The equation demonstrates that density increases with depth and this difference between the hot and cold fluids is a major contributor to the driving force of the convective motion.

As density and temperature increase with depth, so too does pressure because the weight of the overlying material naturally increases with depth down to the core of the planet. The earth has a three-layer configuration consisting of a crust, mantle and core. Each layer has a slightly varied density and thickness. Pressure values can be calculated for the mantle as a function of depth where the core has a radius (a) and the mantle (a-b). Gravity must also be taken into account because it remains nearly constant to the core-mantle boundary. More accurate estimations of mantle pressures can now be found for differing regimes using equation 1-3:

$$p(r) = \frac{2}{3}\pi G \rho_c^2 (b^2 - r^2) + \frac{2}{3}\pi G \rho_m^2 (a^2 - b^2) + \frac{4}{3}\pi \rho_m G b^3 (\rho_c - \rho_m) \left(\frac{1}{b} - \frac{1}{a} \right) \qquad (1\text{-}3)$$

where the following terms are used:

G = universal gravitational constant
ρ_c = density of core
ρ_m = density of mantle
b = core radius where ($0 \leq r \leq b$)
$a-b$ = mantle radius where ($b \leq r \leq a$)
*a = earth radius

This lengthy derivation calculates pressure of varying mantle depth and contituent density. The determination of pressure values is valuable because certain degrees inhibit or promote certain mantle processes. One of these processes is mantle melting and subsequent upwelling of molten material. Pressures of great magnitude can be calculated for varying depths and planet configurations with the aid of equation 1-3.

The increase in pressure values experienced with depth into the mantle requires an increase in temperature to melt mantle rock. Without the heating the temperature desparity between mantle materials and cooler, overlying rock would not be of a sufficient value to drive mantle convection. The melting occurs however and the behavior and properties of the heating can be accurately modelled using an equation that describes the solidus temperature. The solidus temperature (T) is simply the temperature at which the ascending mantle rock and solidus boundary are equivalent. The intense pressures found in the mantle inhibit the actual melting of rock even at extreme temperatures (900-1800k). As convective processes operate, ascending plastic-like material reaches a height at which pressures are reduced enough to allow for melting because the solidus temperature decreases with decreasing pressure. The equation 1-4 for the solidus temperature is as follows:

$$T(^{\circ}k) = 1600 + 0.13p(MPa) \quad (1-4)$$

where

T = solidus temperature

p = pressure

The solidus equation can be used to locate the depth at which we can predict mantle material to initiate melting. Equation 1-4 illustrates the environmental conditions necessary for mantle material to achieve melting which is ultimately the molten component that forms oceanic crust. This equation allows for accurate interpretation of mantle conditions and insight into the behavior of crustal generation.

Once the solidus temperature has been reached by the upwelling mantle material, density contrasts drive the melted material upward while the cooler material descends deeper into the mantle. Heating experienced by the material causes expansion that results in a "buoyancy" effect driving the molten mass towards the surface. The heat found in the mantle regime necessary to continually drive this convective process comes from two primary sources: the cooling of an initially accreted earth and by the decay of rediogenic isotopes. These high temperatures associated with the earth's interior are mostly generated by the decay of:

uranium isotopes U^{238} ∧ U^{235}, the potassium isotope K^{40}, ∧ the thorium isotope Th^{238}.

The idea of isotope heating is a fairly simple concept. Kinetic energy is the product that is yielded from the alpha and beta and gamma rays given off during decay of the isotopes. In turn, the energy is transformed into heat as the paths of the rays are very short in the vicinity of the radioactive source. The actual amount of heat produced by the decay of these isotopes is limited to the concentration of these elements in the earth. Therefore, the determination of the mean heat flow genration per unit mass (H) can be found using equation 1-5:

$$H = \frac{Q}{M} \quad (1\text{-}5)$$

where

Q = heat flow

M = mass of mantle

Heat dissipation from the earth resulting from its initial accretion has naturally diminished over 4 billion years of cooling. The majority of heat conduction thus comes from radiogenic decay. Present-day totals of heat generation (H) can be more realistically and accurately modelled by calculating the total heat generated by the three main radiogenic isotopes since they account for the majority of the heat presently being produced. Heat flow can be found by utilizing equation 1-6:

$$H_o = C_o^u \left[H^u + \frac{C_o^{Th}}{C_o^u} \times H^{Th} + \frac{C_o^k}{C_o^u} \times H^k \right] \qquad 1\text{-}6$$

where the following notation is used:

C_o^u, C_o^{Th}, C_o^k = *present mass concentrations of K - Th- U respectively*
H^U, H^K, H^{Th} = *heat release values*

After analysis of the above equation, individual concentrations of each isotope can be found if the other parameters have been determined. This idea is a very effective tool in modeling mantle heat and isotope concentration behaviors. Typical amounts of these heat-generating elements have been found for many rock types which then enables the geophysicist to determine heat-production values in mantle material of varying mineralogic and chemical composition. For example, if a certain region of the mantle under study has large amounts of tholeiitic basalt in its composition, then if the concentration of the heat producing isotopes in the basalt are known, ultimately, the rate of heat generation can be found for that region of the mantle.

The values of varying heat flow have a direct influence on viscosities found in the mantle. The viscosity parameter is important in the understanding of actual thermal circulation processes. Viscosity measures a material's resistance to change its form or the material's "internal friction". A direct consequence of an ancient accreted earth is that the heat produced by the process lowered the viscosity to a point where convection took over as the primary mechanism for the transport of the heat (Press, 1982). Since depth would take into account both temperature and pressure constraints in calculating mantle viscosities, its value can be modelled using equation 1-7:

$$\mu \propto \exp \frac{\{ (E_a + p \times V_a) \}}{R \times T} \qquad (1-7)$$

where the following notation is used:

E_a = *activation energy (for specific compositions)*
p = *pressure*
V_a = *activation energy per mole*
R = *gas constant*
T = *temperature*

Fluidity of molten material is necessary for thermal convection because it is the transport of heat through the motion of a medium rather than through conduction. The amount of viscosity a material possesses will determine velocities and forces of the mantle circulating cell.

The result of calculating viscosity leads to the determination of the last description of mantle properties to be examined. The value of a material's viscosity affects the amount of stress needed for a mantle convection cell to operate. The actual cell of molten material flows as one would expect, in a seemingly rectangular fashion. As heat is supplied and the viscosity is low enough to facilitate movement of material, the velocity gradient reaches a maximum in opposite corners of the cell (* see solution to equation 1-8 in Appendix A for diagram). With the constraints on viscosity already determinded, the stress required to drive the circulating cell is given by

$$\sigma = \mu \frac{\Delta V}{\Delta Z} \quad (1\text{-}8)$$

where:

μ = *viscosity*
ΔV = *velocity of material*
ΔZ = *depth*
* *the ratio of* $V \wedge Z$ = *the velocity gradient* ϵ *one section of the cell*

This derivation gives a value for the amount of stress that is overcome in order to sustain a convection cell in the asthenosphere. The bouyancy stress needed to drive the upward plume would seemingly be an incredibly high value in order to overcome the viscous mantle melt. Under certain conditions, the amount of stress needed to sustain the cell is comparable to the amount of stress needed to fill an ordinary truck tire with air.

ACKNOWLEDGEMENTS

Equations found in Chapter 2 are taken from Geodynamics. Applications of Continuum Physics To Geological Problems. by Turcotte & Schubert. Ch. 1: and from material prepared by Dr. Hallan Noltimier for class discussions. Example solutions to equations 2-1 through 2-12 are given in Appendix B.

Chapter II

MAGNETICS

OVERVIEW

The concept of paleomagnetometry had its origin in the late 1950's when magnetic anomalies (variations in the magnetic field above and below the accepted value) were discovered in strips of oceanic crust paralleling mid-ocean ridge systems. These anomalies allowed spreading rates and relative plate velocities to be calculated. Values given to these questions allowed geophysicists to determine ancient north poles (paleopoles) of continental masses. This chapter will attempt to describe the magnetization process and solve several problems dealing with locations and velocities of lithospheric plates. The determination of such questions are important but the processes in which the numbers, used in the calculations, are found are needed to accurately model plate tectonics. The steps in reducing experimental error are major concerns if data is to be interpreted accurately. Without an understanding of the values and the error introduced into the calculations, the answer to related geologic problems will have no meaning. As a result, a general description of the error-reduction processes of magnetic calculations, called Fisher statistics, will be examined. An identification and explanation of the major equations and derivations reducing error values accompanying data collection processes will also be given.

STATISTICAL ANALYSIS

As lithospheric plates move across the earth, cracks form and grow as two plates slowly drift away from one another. Hot mantle rock, driven by the thermal convection cell modelled previously, flows upward to fill the gap caused by the spreading plates. This drifting and upwelling of new material occurs at mid-ocean ridge systems where new oceanic crust is continually being generated. The molten material is extruded from the ridge and begins to cool, crystallize as it spreads away from the crest. The process of magnetization begins at the point of extrusion from the magma chamber and subsequent crystallization at the ridge.

The dominating mineral found in the mantle is the silicate mineral olivine. This mineral contains iron which is ferromagnetic and can acquire a permanent magnetism from the earth's natural magnetic field when the material is formed. As the molten rock cools, it acquires a thermoremanent magnetism (TRM). This magnetization process involves the magnetic moment in ferromagnetic substances and in the spins of its electrons. This section will concentrate on the applications of the magnetic data and its compilation and not on the specifics of the magnetization process. A rock, upon magnetization, can now be analyzed to determine its orientation and direction at the time of its formation.

In order to accurately interpret data collected from magnetized rocks, several steps must be followed to eliminate errors introduced during the process. First, to estimate a sample's paleolocation, it must be determined if the rock's present magnetization is its initial one and not an overprinting of another magnetization which it may have acquired since its formation. This question can be satisfactorily answered using field relations, mineralogy and heating the sample in the laboratory to eliminate (thermal demagnetization) post-crystallization acquisition of magnetism.

The next step in reducing magnetic data is in the collection of samples for analysis. The declination and inclination of each rock core taken must be measured very carefully to orient the rock in the true direction. The declination (D) is the angle between geographic north and the magnetic field direction measured from 0 to 360 degrees. The inclination (I) is the angle between the horizontal and the field direction measured downward from +90 degrees to -90 degrees. To accurately determine the results of a number of samples, a process known as Fisher statistics can be used to sensibly average all the data collected. This analysis is used primarily in paleomagnetism to average all sorts of vectors. Fisher statistics also yields several other parameters including the cone of confidence which will be described later. After the locations have been logged, the direction of the remanent magnetization vector (J) must be calculated. The direction of vector (J) can be described using a very simple coordinate system that relates the northern, eastern and downward components of (J) [* see related solution to equation 2-1 in Appendix B for diagram]. The relations are given by the following values

$$\begin{aligned} \alpha &= \sin I \\ \beta &= \cos D \cos I \\ \gamma &= \sin D \cos I \end{aligned} \qquad (2\text{-}1)$$

α being the easterly component of vector J
β being the northern component
γ being the downward component
D = declination of collected sample
I = inclination

The sums of all three directions for all of the collected samples are then used to find the length of the resultant vector (R) which is simply another notation for the magnetic vector (J). (R) is found using equation 2-2.

$$\vec{R} = \sqrt{(\sum\alpha)^2 + (\sum\beta)^2 + (\sum\gamma)^2} \qquad (2\text{-}2)$$

where:

$\sum\alpha$, $\sum\beta$, $\sum\gamma$ = *the total* $\sum$ *of each directional component*

Equation 2-2 calculates the length (R) of the single remanent magnetic vector (J) with respect to alpha, beta and gamma. The individual directions that comprised vector (J) can be found for the derived total length of (R), given by equation 2-3:

$$\frac{\sum\alpha}{R}, \frac{\sum\beta}{R}, \frac{\sum\gamma}{R} \qquad (2\text{-}3)$$

From these values, the inclination and declination of the initial magnetization field that the sample acquired can be calculated using equation 2-4:

$$I_R = \sin^{-1} \frac{(\sum \alpha)}{R} \qquad D_R = \tan^{-1} \frac{(\sum \alpha)}{\sum \beta} \tag{2-4}$$

This error-reducing process of determining the actual magnetization direction of a rock during its formation can help in the location and orientation of the rock when it initially crystallized.

The coordinates used to describe locations of the earth are given as the latitude and longitude. Latitude is the north, south component measured from the equator and longitude is the east, west component measured from Greenwich, England. To pinpoint the initial location of the sample would be to find its paleolatitude and paleolongitude. These two directions give the location of the magnetic north pole (paleomagnetic north) at the time the rock formed. If the declination and inclination of the remanent magnetism is known, then with the coordinates recorded for the sampling site the paleomagnetic directions can be found. The first step in deriving an answer to this is visualizing the relative positions of the poles and calculating the angular distance

V θm , *the magnetic colatitude,*

between the sampling site and the position of the paleomagnetic north pole. The position of the paleomagnetic pole is not needed here because the colatitude is related to the inclination of the site with respect to the pole. The geographic north pole, the paleomagnetic north pole and the sampling site create a spherical triangle in which the relative positions of all the poles can be calculated [* see solution to equation 2-5 in Appendix B for diagram]. The equation for the magnetic colatitude is given by:

$$\tan I = 2 \cot \theta m \qquad (2\text{-}5)$$

where:

I = *remanent inclination*
θm = *colatitude*
$0^\circ \leq \theta m \leq 180^\circ$

This is the distance of one side of the spherical triangle for which other values can now be calculated using spherical geometry. With the remanent (I) and (D), the latitude and longitude of the sampling site and the colatitude, the paleolatitude can be derived using equation 2-6:

$$\sin\Phi p = \sin\Phi \cos\theta m + \cos\Phi \sin\Phi \cos D \qquad (2\text{-}6)$$

where the following notation is used:

$$\Phi = \textit{site latitude}$$
$$\theta m = \textit{colatitude}$$
$$D = \textit{remanent declination}$$

The paleolongitude can then be calculated if the value for the paleolatitude is first checked. The reason for this procedure is that by going from the declination to inclination in the calculations, there is a possibility of being off 180 degrees in the measurement. Two test formulas can be used to determine the method needed to derive the paleolongitude. The first is equation 2-7 and is given by:

$$\cos\theta m > \sin\phi \sin\phi p \qquad (2\text{-}7)$$

where:

$$\phi p = \textit{paleolatitude}$$

If this equation is correct and the colatitude is greater, then equation 2-9 can be used to calculate the paleolongitude:

$$\sin(\psi p - \psi) = \frac{\sin\theta m \sin D}{\cos\phi p} \qquad (2\text{-}8)$$

where the following definitions of terms are used:

$$\psi p = \textit{paleolongitude}$$
$$\psi = \textit{site longitude}$$
$$D = \textit{remanent declination}$$
$$\phi p = \textit{paleolatitude}$$

Equation 2-8 derives the paleolongitude of the paleomagnetic north pole measured from the sample's location and orientation. If however the cosine of the colatitude makes equation 2-7 false, then equation 2-9 must be used to compensate for the 180 degree rotation experienced by going from the declination to the inclination:

$$\sin\left(\pi + \psi - \psi p\right) = \frac{\sin\theta m \sin D}{\cos\phi p} \qquad (2\text{-}9)$$

where the following notation is used:

ψ = *site longitude*
ψp = *paleolongitude*
ϕp = *paleolatitude*

Equations 2-6, 2-8 and 2-9 allow the paleo-coordinates for the paleomagnetic pole position to be calculated at the time when the sample was initially crystallized. If the error reduction process and the sampling procedures were completed efficiently, the values that were found should accurately represent the position of the magnetic pole seen in the sampling cores. The same concept used in the calculations of ancient magnetic poles can be used to determine the velocities and rates of spreading between mid-ocean ridge systems and between lithospheric plates.

The magnetic anomalies preserved in strips of oceanic crust are a result of variations in the earth's magnetic field throughout history. As the igneous rocks, extruded at the ocean ridges, cool they acquire a thermal remanent magnetism in the direction of the earth's present field. The widths of the magnetic strips can be used to determine the velocity of the seafloor spreading. The distance from the ridge crest to each anomaly is plotted against the known time of each episodic reversal in the magnetic field. The slope of this line, given by equation 2-10, will give the rate of seafloor spreading:

$$M\ (slope) = \frac{Y_2 - Y_1}{X_2 - X_1} \qquad (2\text{-}10)$$

The magnetic anomaly profile pattern and the correlation graph of age verses time gives the slope which in turn gives the velocity of the oceanic crust moving away from the mid-ocean ridge.

The spreading velocities of the lithospheric plates can also be calculated using geometric expressions. The surface area of the earth remains essentially constant so the velocities of seafloor spreading can be related to the rate of crust subduction at oceanic trenches. As a result, relative velocites between rigid plates can be determined. A rigid surface plate can be translated to a new position by rotation about a uniquely defined axis. This motion between two plates is completely described by the latitude and longitude of the axis of rotation and the angular velocity. The velocity then can be calculated for the rate of spreading between two adjacent plates given by equation 2-11.

$$u = \omega\, a \sin \Delta \qquad (2\text{-}11)$$

where the following definitions of terms are used:

ω = angular velocity of rotation
a = radius of the earth
Δ = the $\angle$ subtended at the center of the earth by the coordinates of the axis of rotation Λ a reference point on the plate boundary

The angle is a component of the spherical triangle which comprises the colatitude of the east longitude of both the pole of rotation and plates boundary coordinates. The angle can be found using equation 2-12: [* see solution to equations 2-11 and 2-12 in Appendix B for diagram].

$$\cos\Delta = \cos\theta\cos\theta' + \sin\theta\sin\theta'\cos(\psi - \psi') \qquad (2\text{-}12)$$

where the following terms are used:

θ = colatitude of rotation axis
θ' = colatitude of reference point on plate boundary
ψ = east longitude of rotation axis
ψ' = east longitude of reference point

Equations 2-11 and 2-12 derive the spreading rates and lithospheric plate velocities in constant dynamic action on the earth's surface.

Chapter III

DISCUSSION

The primary objective of the field of geophysics is to develop a thorough and cohesive model for the dynamic Earth system encompassing the essentials of the planet's thermal, chemical and physical structure. Through the use of classical derivations, numerical values can be placed on fundamental geologic processes that help in the understanding of plate tectonics and a variety of natural phenomena. All of the expressions in this paper are presented with the minimum of complexity so that a general yet thorough modelling of certain processes can be achieved. Mathematical problems that arise are vital in grasping the underlying principles of geological activity so example solutions have been presented in Appendix A and B.

Mantle processes, the key to many major plate tectonic phenomena, are not easily accessible to scientific study. The synthesis of simple expressions to describe mantle properties are extremely useful and very accurate. The basics of any geologic study is knowing the general environment of the subject under analysis. For the mantle, density, pressure and temperature are the three main parameters that have been modelled. Variations of these parameters can also be calculated by the addition of new constraints into the expression: for example, changes with depth and pressure have a significant effect on the temperature and velocity of the mantle. Thermal properties of mantle regions can be found knowing the heat-producing elements and the mass of the material being heated. In turn, heat flow can be modelled for areas of the mantle with varying composition. Because every parameter has some degree of effect on the others, expressions can be modified to adjust to changes in their value. The viscosity takes into account the temperature and pressure of the mantle and the stress needed to drive the mantle thermal cell has viscosity as one of its constraints. A simplified yet thorough introduction of mantle processes and properties are given by equations 1-1 through 1-8 which describe pressure, density, pressure changes with depth, temperatures of mantle melting, heat flow, heat flow of individual isotopes, viscosity and finally stress.

Chapter two deals with describing major plate movements and the reduction of data to improve the experimental accuracy. Magnetic data which is a relatively new concept has provided a wealth of knowledge. The magnetic anomaly patterns seen in oceanic crust has provided a tool for calculating spreading rates which can then be related to velocities of major lithospheric plates. In order to pinpoint actual locations and orientations of crystallized rocks, their magnetic character must be accurately described. Fisher statistics is a useful tool in reducing magnetic error and involves the simple synthesis of vector analysis. Spherical geometry is then used to transform the magnetic data into actual coordinates of paleolatitude and paleolongitude. By using the magnetic character of the rocks, velocities and spreading rates can be calculated. The accuracy of any expression is increased with the reliability of the numbers put into it. By utilizing Fisher statistics the orientations and velocities of the earth's surficial plates can be accurately determined. These expressions help to solve the puzzles of plate interactions that produce subduction, the opening and closing of oceans, continental collisions, the generation of crust and host of other processes.

The earth is a very dynamic system that is endlessly changing shape, composition and character. New methods of thought and scientific research must be sought in order to keep in touch with our planet and the neighboring galaxies. The geophysical approach in modelling earth processes is just one method of the many disciplines combining chemistry, physics and mathematics to describe the earth. Where many parts of the earth are inaccessible or tools have not been developed to directly measure certain earth regimes, geophysical expressions can play an integral part in initiating studies and analyzing the regimes under different constraints. The ultimate goal of any modelling attempt is to construct a planet that would include accretion, melting, crystallization and translation of materials. As the field of geophysics expands many expressions will be derived to answer the problems encountered during modelling and hopefully will initiate new studies to fully explain the earth and its processes.